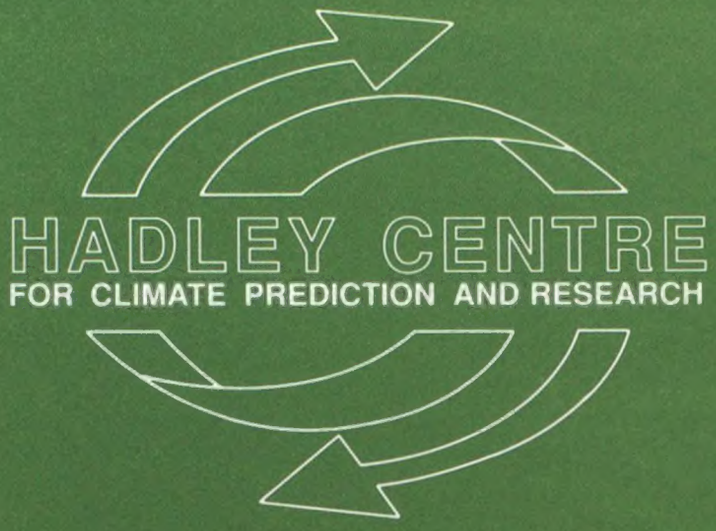

Spinup methods for HadCM3L

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C D Jones and J R Palmer

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April 1998

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Hadley Centre
Meteorological Office
London Road
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The Met Office

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SPINUP METHODS FOR HADCM3L

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Spinup methods for HadCM3L

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April 9, 1998

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Abstract

This note describes a new method for spinning up an ocean-atmosphere coupled general circulation model. The ocean component of a coupled model takes much longer to reach a near-equilibrium state than the atmospheric component does. The spinup method described here involves using climatic data from the atmospheric component of a coupled model run to drive a long ocean only run with a lengthened timestep. By allowing much of the ocean spinup to occur in an accelerated ocean only phase, a significant saving of computer resources can be achieved. Results are presented which show that the coupled model can be satisfactorily spun-up with a saving of almost 95% in computer time.

1 Introduction

General circulation models (GCMs) are currently the best tools available for understanding the current climate and predicting future changes. However, they typically show a climate drift when simulations are run from initial conditions created from climatological estimates of the current states of the atmosphere and ocean. Climate change experiments using GCMs need to start close to the model's equilibrium so that the drift does not dominate over the signal of climate change. The process of achieving this equilibrium state is know as "spinup".

The obvious way to achieve a spun-up initial state is to begin from an estimate of the current climate and run a model from there until the drift is small enough to be acceptable. This was the approach adopted in the HadCM2 project (Johns et al. 1997). This method is expensive in terms of computer time (typically several hundred years of simulation are required). Because the equilibration timescale of the ocean is much longer than that of the atmosphere (hundreds of years compared with several years), it would seem sensible to treat the two components of a coupled GCM separately. However, interactions and feedbacks between the two components ensure that the coupled equilibrium is not equivalent to the two independent equilibria, and a drift still occurs when the two components are recoupled.

This report describes a new faster method to spinup a coupled climate GCM. The method was developed for use with HadCM3L which is a version of the third Hadley Centre Climate Model, HadCM3 (Gordon et al. 1998). This model is being used within the Coupled Climate-Carbon Cycle Project (CCC Project, Cox 1997a) which aims to include a fully interactive carbon cycle in a climate change experiment. The theory of periodically synchronous coupling techniques for speeding up model evolution is presented in section 2. Section 3 describes HadCM3L, concentrating on the differences between it and HadCM3, and outlining the additional requirements placed upon it by the carbon cycle sub-models which necessitate the use of flux adjustments. The application of such a technique to the spinup of HadCM3L is discussed in section 4, and problems encountered during an initial experiment are discussed in section 5, along with their solutions. Results from a second spinup experiment are presented in section 6. Section 7 draws some conclusions.

2 Asynchronous coupling

The idea of using asynchronous techniques to speed up the evolution of a coupled model is not new, but applying such ideas to a GCM is not yet widespread. Most studies to date have focussed on simple (often linear) energy balance "box" models.

Before discussing the application of such techniques, some terminology should first be defined. Synchronous (SYN) running of a coupled model means that the atmosphere and ocean components are integrated quasi-simultaneously with frequent exchange of information. This is the normal mode of running a cou-

pled model. For example a typical coupled GCM may be run for one day in atmosphere mode, and then mean forcing fields are used to run for one day in ocean mode. The state of the ocean is then used as a boundary condition for the next atmosphere day, and so on. Asynchronous (ASYN) running is when the components are run alternately for long periods. For example the atmosphere could be run for a year, and then mean fluxes used to run the ocean for 10 years. Strictly speaking synchronous running is simply asynchronous running with a short coupling period.

Clearly, asynchronous techniques would save a lot of computing time if they could accurately reproduce the evolution of synchronously run models. This is especially true if the atmosphere component of the model is much more expensive computationally than the ocean component (which is often the case), and if long ocean-only phases can be used without loss of accuracy. Sausen and Voss (1996) demonstrated that purely asynchronous running of a simple box model could accurately reproduce the evolution of a synchronous run, but only if there was no seasonal cycle in the forcing of the model. When a seasonal cycle was included, the asynchronous approach could not cope with the changes in fluxes during the cycle.

To combat this problem they used the approach suggested by Gates (Schlesinger, 1979) which instead of running with an "atmos-ocean-atmos-ocean" type cycle, runs with a "coupled AO-ocean only-coupled AO-ocean only" type cycle. This technique is called periodically synchronous (P-SYN) running, and Sausen and Voss (1996) showed that it performed much better than the ASYN technique when faced with a seasonal cycle of forcing.

The simple box models of Sausen and Voss do not represent the high degree of internal variability present in a coupled AOGCM. Voss and Sausen (1996) went on to show that when white noise variability was included in the simple models, even the P-SYN technique could fail. The reason for this is that the noise caused variability in the coupled phase of the run and so non-representative periods may then be used to determine the fluxes to be applied to the long ocean-only phase, causing incorrect drift of the ocean state. The solution to this problem was found to lie in using a weighted mean of fluxes from the most recent coupled phase and from earlier phases.

Roberts (1990) and Davey (1991) tested the suggestion that the derived fluxes should evolve as the ocean only phase evolved, rather than being held constant. They compared explicit and implicit forcing methods. The explicit forcing method used the atmosphere and ocean states from the last coupled phase to determine the fluxes, which were then held constant for the duration of the next ocean-only phase. The implicit forcing method involved determining fluxes using the atmospheric state from the last coupled phase together with the current ocean state from the ocean only phase. In this way, the fluxes evolve with the ocean. Fully explicit methods can support oscillations of the ocean state if it adjusts too far to the constant forcing during the ocean only phase. Fully implicit methods do not allow for enough forcing from the atmosphere, and convergence is very slow. Both studies found that the best method was a semi-implicit approach made up from a linear combination of both methods. A

weighting of 95% explicit, 5% implicit was found to be optimal for their linear box model.

Simple box models can be used to decide what coupling techniques are likely to fail, but cannot guarantee success when applied to a full GCM. Other considerations are whether it is practicable to apply them. For example the semi-implicit approach described above was simple to implement and tune in a box model, but as Roberts (1990) notes, the optimal weighting of explicit and implicit forcing in a GCM is model dependent and will vary in time and space, making it very complicated to implement. Also, in calculating the implicit forcing, it would be necessary to run a lot of the physics of the atmosphere model with the new ocean state. This would therefore make the ocean only phase almost as expensive as the full coupled phase.

Sausen et al. (1997) extended the techniques presented in Sausen and Voss (1996) and Voss and Sausen (1996) to apply a P-SYN approach to a coupled AOGCM. Results showed that the approach was successful, with P-SYN runs accurately reproducing both the equilibrium state and transient behaviour of the synchronous run of the GCM to changes in radiative forcing, at a saving of about 70% CPU time. Here we are faced with a simpler task. We are not trying to simulate climate change, but merely bring a coupled GCM to equilibrium. Thus it seems likely that a P-SYN approach will be successful; if not in exactly creating an equilibrium state, then at least in bringing the model close enough to it that only a short period of coupled running will be required subsequently.

3 Model set up

The coupled atmosphere-ocean GCM used for this study was a version of HadCM3, but with a low resolution ocean component in order to save on computer time. The ocean resolution was 2.5° x 3.75° (the same as the atmosphere component) rather than 1.25° x 1.25° as in HadCM3. This low resolution model is referred to as HadCM3L. The model is a configuration of the UKMO Unified Model (UM, Cullen 1993). The atmosphere component is a 19 level quasi-hydrostatic grid-point model based on the full primitive equations on an Arakawa "B" grid. The ocean component has a grid congruent with the atmosphere grid to facilitate coupling. It is based on the primitive equation model of Bryan (1969) and Cox (1984), with 20 levels in the vertical, and a realistic topography. A zero-layer thermodynamic sea-ice model is also present, on the same grid, including ice and snow advection. A more detailed description of the model and some model specific parameters are included in Appendix A.

There are some minor differences in model formulation between HadCM3L and HadCM3. HadCM3L has a spatially constant coefficient for the Gent-McWilliams eddy parameterisation scheme (Gent and McWilliams, 1990). The Mediterranean outflow parameterisation is grid dependent and has been modified slightly. HadCM3L, like HadCM3, runs with a half hour timestep for the atmosphere component, and a one hour timestep for the ocean component. The coupling period is 1 day.

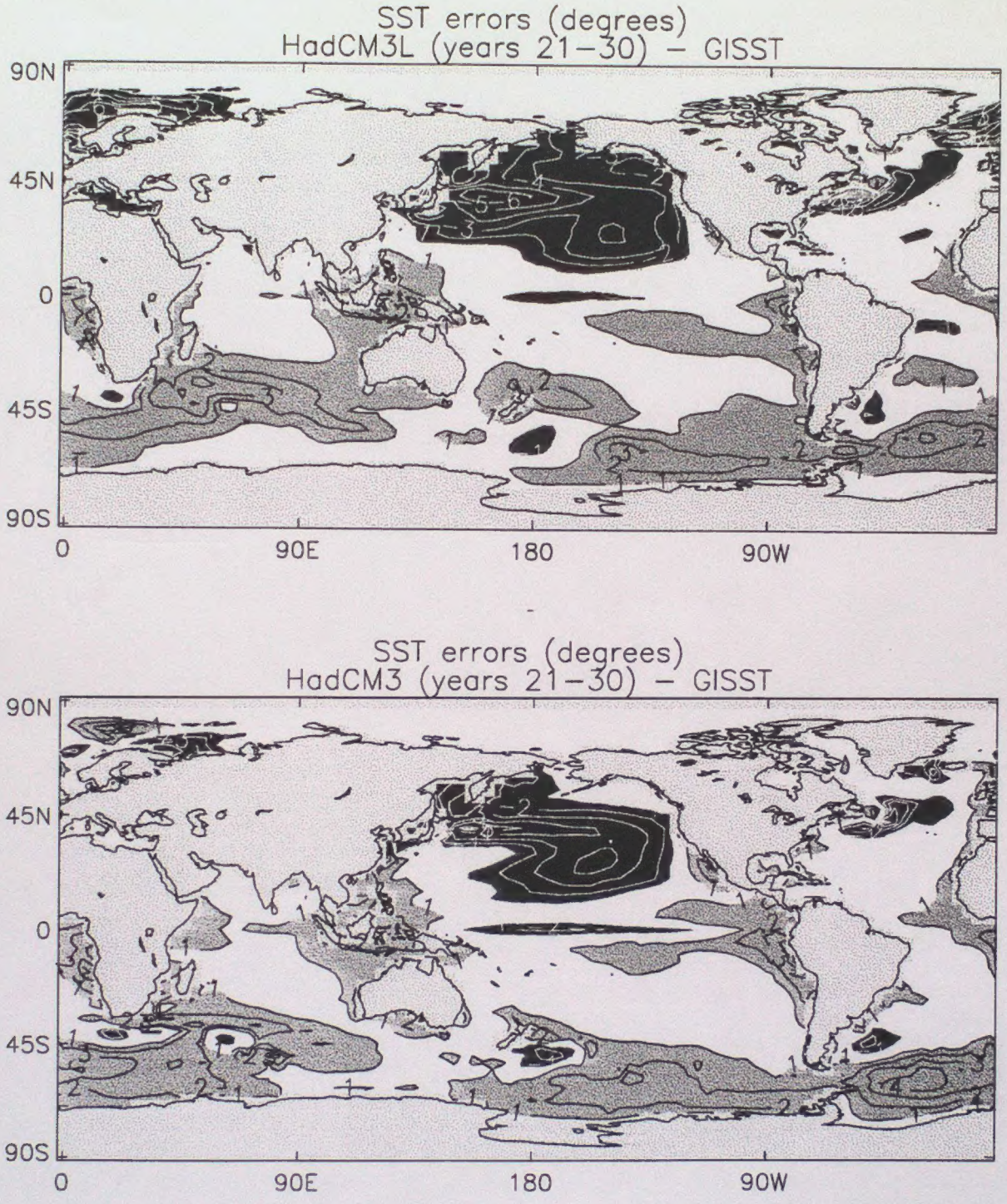

Figure 1. Decadal mean sea surface temperature errors relative to GISST climatology, averaged over years 21-30 of unflux-adjusted HadCM3L (top panel) and HadCM3 (bottom panel).

A major difference between the running of HadCM3L and HadCM3 is in the use of flux adjustments. Whereas the HadCM3 model has been run successfully without any flux adjustments, in an un-flux adjusted trial of HadCM3L some serious systematic errors were observed after running for forty years. Some of these were also present in HadCM3, but others were the result of the lower ocean resolution.

Figure 1 shows the decadal mean distribution of sea surface temperature (SST) errors, relative to GISST (Global sea-Ice and Sea Surface Temperature data set, Rayner et al. 1995), averaged over years 21 to 30, for both HadCM3L (top panel) and HadCM3 (bottom panel). Apart from a large cold error in the North Atlantic which is due to a sea-ice problem described below, there is a large cold pool in the central Pacific, and a warm band in the Southern Ocean. This error pattern is very similar in magnitude and appearance to that seen in HadCM3. There was also seen to be a long term drift in volume weighted salinity (fig 2), due mainly to the formulation of the P-E calculation (the same drift was seen in HadCM3).

However, due to the coarser resolution, HadCM3L was not able to sufficiently represent the gulf stream, and hence the northward transport of warm water associated with it. This led to a gradual build up of sea ice in the North Atlantic, and the cold pool seen in the SSTs. Figure 3 shows the extent of polar sea ice in March (when it is at a maximum in the northern hemisphere) for year 30 from HadCM3L (top left panel), HadCM3 (top right), and the GISST climatology (bottom left). It is clear that HadCM3L has ice much too far south into the Atlantic. HadCM3 performed much better, and kept a realistic ice distribution. The consequence of this excess of ice is that the formation of North Atlantic Deep Water is hindered and the Atlantic overturning streamfunction (not shown) is too weak.

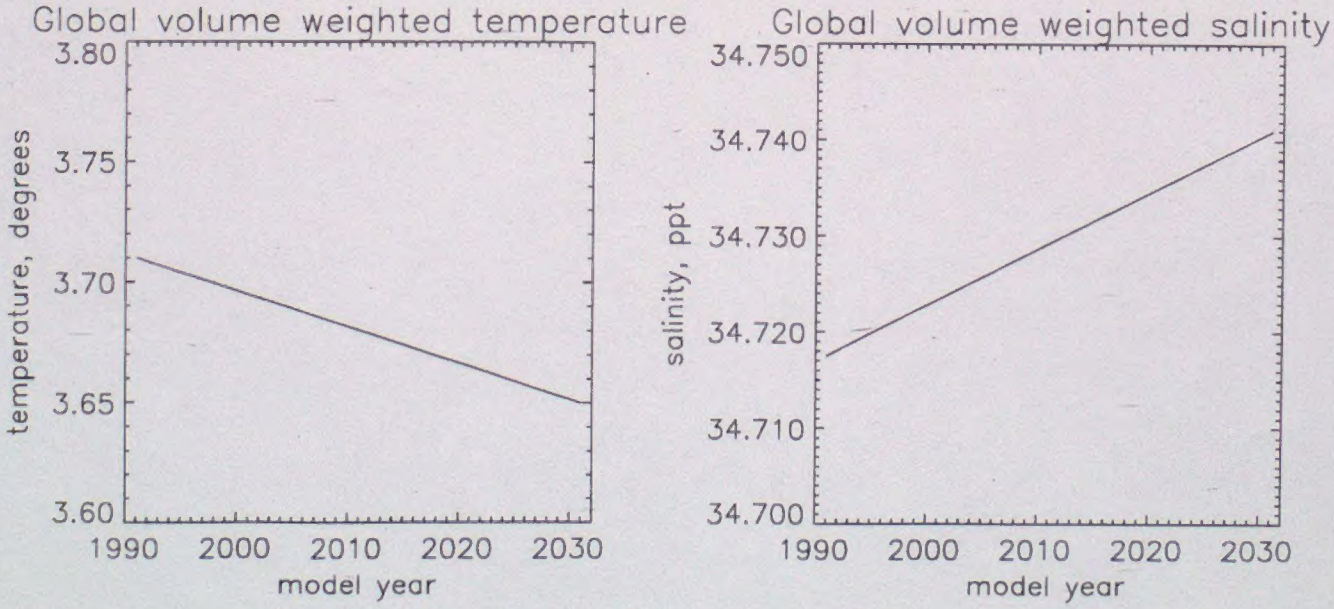

Figure 2. Global, volume weighted, mean temperature and salinity during the 40 year unflux-adjusted test of HadCM3L.

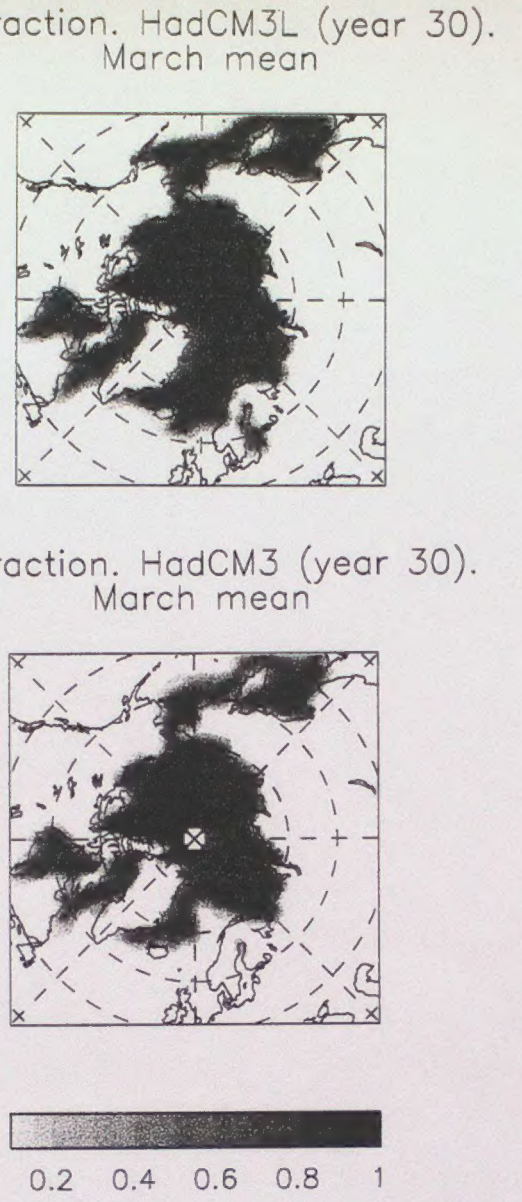

Figure 3. Northern Hemisphere maximum extent sea ice fraction from HadCM3L, year 30 (top panel), HadCM3, year 30 (middle panel) and GISST Climatology (bottom panel).

It is important to remember that the aim of the coupled model in the coupled carbon cycle project is to provide an environment within which to run the carbon cycle model. Two sub models, HadOCC (Hadley Centre Ocean Carbon Cycle, Palmer and Totterdell, 1998) and TRIFFID (Cox 1997b) aim to represent the key chemical and biological processes of the carbon cycle in the ocean and terrestrial biosphere respectively. It was felt that although the errors in HadCM3L were not so severe as to prevent a useful climate study being performed with the model, they were significant enough to affect the results from the carbon models. For example a lack of precipitation led to deforestation in India and the Amazon, and the systematic errors in the ocean temperature and salinity fields led to errors in the evolution of the ocean biology. Therefore it was decided that flux adjustment was necessary to keep the HadCM3L ocean surface temperature and salinity realistic.

4 Spinup technique

The spinup methodology used for the second Hadley Centre coupled climate model, HadCM2, was to run the model synchronously coupled for about 500 years (Johns et al. 1997), with alternating Haney forced and flux adjusted phases (see Appendix A for a description of Haney forcing and flux adjustments). The computing cost of using such a spinup method is very high, and would become prohibitive because the carbon fields within HadOCC take several thousand model years to reach equilibrium. There is also a cost associated with the inclusion of additional tracers within the ocean model. Therefore a new method was required that was both cheaper in terms of computing time, and quicker in terms of real time.

The method devised for this purpose is based on a periodically synchronous technique (as described in section 2). The fact that the time scales for drift in the different components of the model differ significantly (the ocean adjusts on timescales of hundreds or even thousands of years, whereas the atmosphere adjusts over periods of years) means that a long ocean-only phase is appropriate.

The method proposed for the spinup is shown schematically in figure 4. In the first stage the coupled model is run with Haney forcing boundary conditions for 60 years, from a starting point based on climatological temperature and salinity data for September from Levitus (1994a,b). A temperature Haney forcing coefficient of 163.76 $Wm^{-2}K^{-1}$ is used, corresponding to a relaxation time of about 2 weeks (for a 50m mixed layer), with a coefficient for salinity relaxation corresponding to the same time scale.

The second stage is a long run of 2000 years in ocean-only mode using the distorted physics technique (Wood, 1998), based on that of Bryan (1984), which allows a lengthened timestep of 24 hours to be used. This means that a long period can be run quickly and cheaply, allowing the deep ocean structure to come close to equilibrium. Climatological temperature and salinity data are again used to Haney force the run with the same coefficients as before. Other fluxes are derived from monthly mean values calculated from the last ten years of the cou-

pled phase. Using monthly means from ten years of the coupled model reduces the risk that the forcing for the ocean only phase will be non-representative of the mean forcing seen by the ocean component during the coupled phase. The forcing fluxes used are penetrative solar heat flux, non-penetrative heat flux, wind mixing energy, wind stress (N-S and E-W components) and net fresh water flux (precipitation minus evaporation). The fluxes are calculated to include the effects of the ice model (Appendix B gives technical details of how this was done in the UM) and thus the need to include the ice model in the ocean only phase is eliminated.

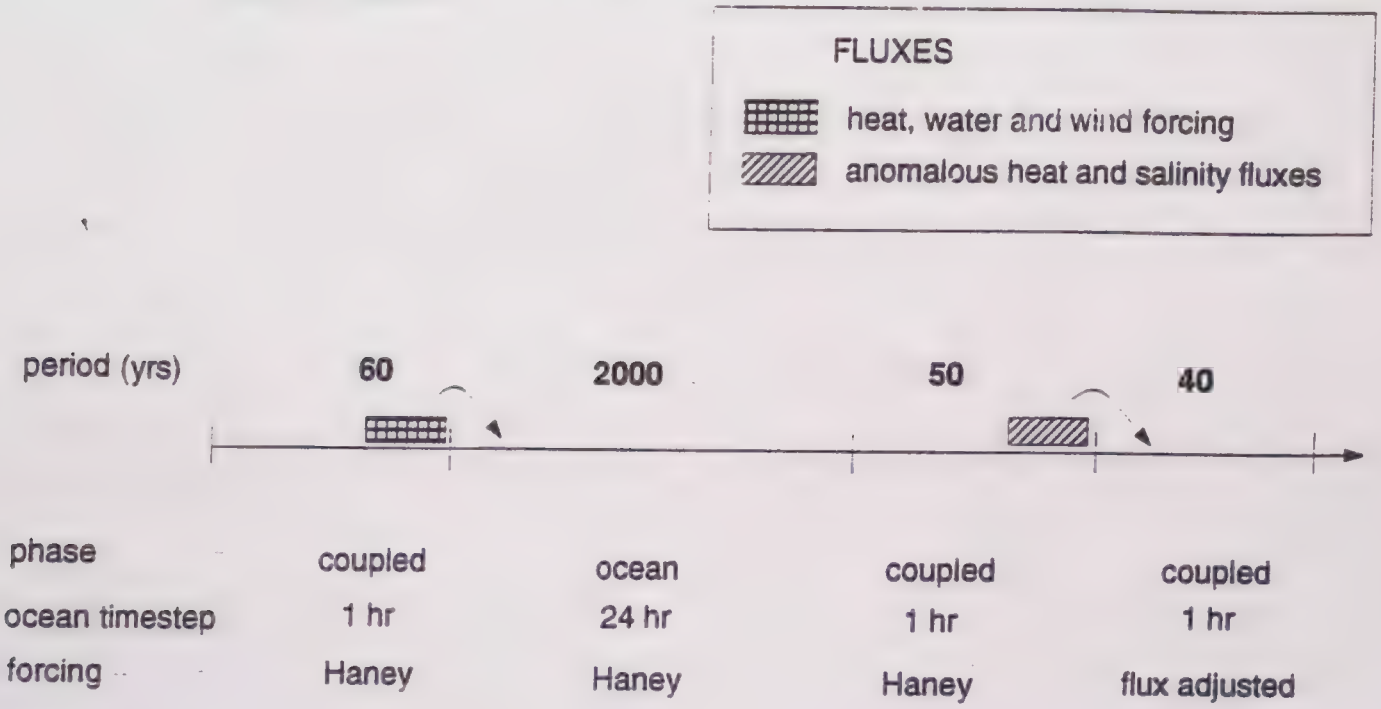

Figure 4. Schematic diagram showing the different phases of the spinup technique and the coupling between them.

In the third stage the ocean is recoupled to the atmosphere model, the ocean state from the end of the ocean only phase being used with the atmosphere state from the end of the first phase. This is run for a further 50 years, again with the same Haney forced boundary conditions. This strategy differs from those outlined in section 2 in two main ways. Firstly, only one ocean-only phase is employed. This phase is much longer than the individual multiple phases used by Voss et al. (1997). Secondly, the ocean model has been accelerated in the ocean only phase by the use of an extended timestep. Previous studies maintained the same model set-up between phases.

To test whether or not this technique produced a model state near equilibrium, a 40 year run using flux adjustments was performed, with the anomalous

temperature and salinity fluxes being diagnosed as 5 day means from the final 10 years of the second coupled phase. The results are discussed in sections 5 and 6.

The real time taken for the spinup (excluding the flux adjusted phase) was about 7 weeks, consisting of roughly 3 weeks for each coupled phase (about 20 model years per week), and 1 week for the ocean only phase. The computing cost of the 2000 year accelerated ocean phase was equivalent to approximately 20 modelled years of coupled running. The spinup as a whole therefore achieved 2110 years of model time in the CPU time equivalent to 130 years of fully coupled running, thus saving almost 95% of the cost.

5 MLD dependence on forcing

A spinup experiment was performed using the plan described in section 3. To determine whether or not the spinup was successful, it is useful to examine the drift of mean ocean temperature and salinity. If the model is near to equilibrium these quantities will be nearly constant. Some inter-annual variability is to be expected, but the long term mean should show no persistent drift. Figure 5 shows the evolution of global volume weighted ocean temperature and salinity (note the irregular time axis - an equal space has been used for each phase). It can be seen in both cases that there was rapid drift during the first coupled phase, which was reduced significantly by the end of the spinup period, and into the flux adjusted run. The table below lists the mean drift in each period of the spinup. It can be seen that the volume weighted mean ocean temperature in the flux adjusted phase drifts slowly. The quality acceptance criteria specified in the PID (Project Initiation Document, Cox 1997a) for the CCC project is a temperature drift of less than 0.1 °C/100 yrs. Although there is no formal criteria for the salinity drift, it is felt that a value of less than 0.1 ppt/100 yrs would be acceptable. By comparison, the drift during the first 40 years of the unforced HadCM3L run is 0.13 °C/100 yrs, and during the initial phase of HadCM3 it is about 0.1 °C/100 yrs.

Spinup Phase	Temperature (°C / 100 yrs)	Salinity (ppt / 100 yrs)
Coupled, Haney I	- 0.18	- 0.012
Ocean only	- 0.028 (first 1000 yrs) - 0.0012 (last 1000 yrs)	- 0.0013 + 0.0003
Coupled, Haney II	- 0.032	+ 0.0037
Flux Adjusted	- 0.0085	+ 0.0074

The table shows that the temperature and salinity drifts increase once the model is recoupled after the long ocean-only phase. The transition from the ocean-only phase to the recoupled phase was examined more closely to try to understand what was the cause of this "recoupling shock". When the mixed layer depths (MLDs) during the spinup were examined, it was noticed that they were typically very similar in the two coupled phases, but rather different dur-

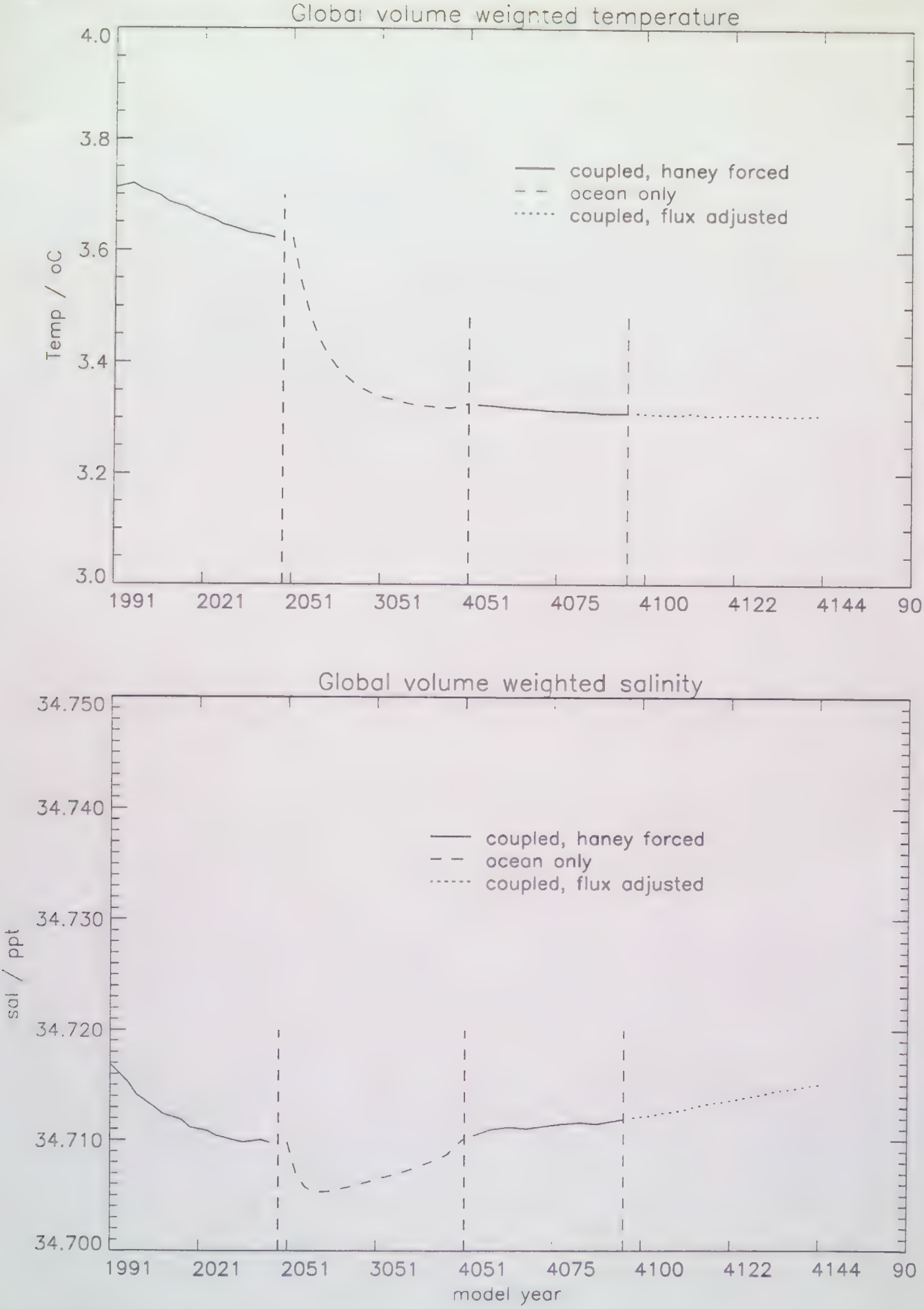

Figure 5. Timeseries of global, volume weighted, temperature (top panel) and salinity (bottom panel) during the course of the spinup. Lengths of each phase are 60 years, 2000 years, 50 years and 44 years respectively.

ing the ocean only phase (most noticeably, they were deeper in September to November in the Southern Ocean). Mean values in the ocean only phase were 40-50% higher than in the coupled phase. Figure 6 shows maximum (top panels), and minimum (bottom panels) MLD values for a single November from the end of the ocean-only phase and the start of the recoupled phase.The maximum values can be seen to agree fairly well (within 20%), but the minimum values in the ocean-only phase are much greater (on average almost double) than the corresponding coupled values.

The reason for this behaviour was found to be the method of forcing the ocean only run with atmospheric parameters taken from the coupled phase. Various heating and mechanical forcings (as described in section 3) are applied as boundary conditions to the ocean. Values of these had been diagnosed as monthly means from the final decade of the first coupled phase of the spinup. There is a strongly non-linear relationship between the mixed layer depth and these parameters (for example the penetration of wind mixing energy drops off exponentially with depth), which means that the response of the ocean to mean forcing is not the same as the mean response to variable forcing. In particular, a decrease in wind mixing energy can cause a large shallowing of the diagnosed MLD, whereas a similar magnitude increase in wind mixing energy would cause a much smaller deepening of the mixed layer. Thus the diagnosed MLDs from the ocean-only phase had realistic maximum values, but minimum values which were much too high, and so were too high on average. This had the consequence that too much heat was being mixed into the mixed layer from the atmosphere (remember that the surface temperature is held constant by the Haney forcing), and so when the model was recoupled, and the MLDs decreased to their previous values, heat was released back into the atmosphere, and the ocean temperature started to drift more quickly.

A solution to this problem is to include daily variability in the forcing data. A rerun of the ocean only phase of the spinup was performed using daily forcing data. It is not desirable to simply use daily data from the final year of the coupled phase because it may be an unrepresentative year in terms of mean forcing fields. So the forcing was calculated by taking the daily values for the final year, and adjusting them by a fixed monthly amount to ensure that the monthly mean values were consistent with the climatology of the final ten years of the run. Equation 1 shows how the daily varying forcing fields, f_{oc}, are calculated from the daily values of the final year of the coupled phase, f_{cp}, shifted by the difference between the monthly means during the final year, F_{year}, and the monthly means during the final decade, F_{decade}. This equation is very similar to the weighting used by Voss and Sausen (eqn. 7 of Voss and Sausen, 1996), but the decadal mean replaces the use of previous coupled phases.

$$f_{oc} = f_{cp} + F_{decade} - F_{year} \tag{1}$$

It should be noted here that the above shifting of daily values was performed at each gridpoint, so that each point maintained its decadal mean for each month. At some points, the difference between F_{decade} and F_{year} was quite large so that some values of f_{oc} became negative. This does not cause a problem for fields such as heat flux which can vary in sign, but it does cause a problem

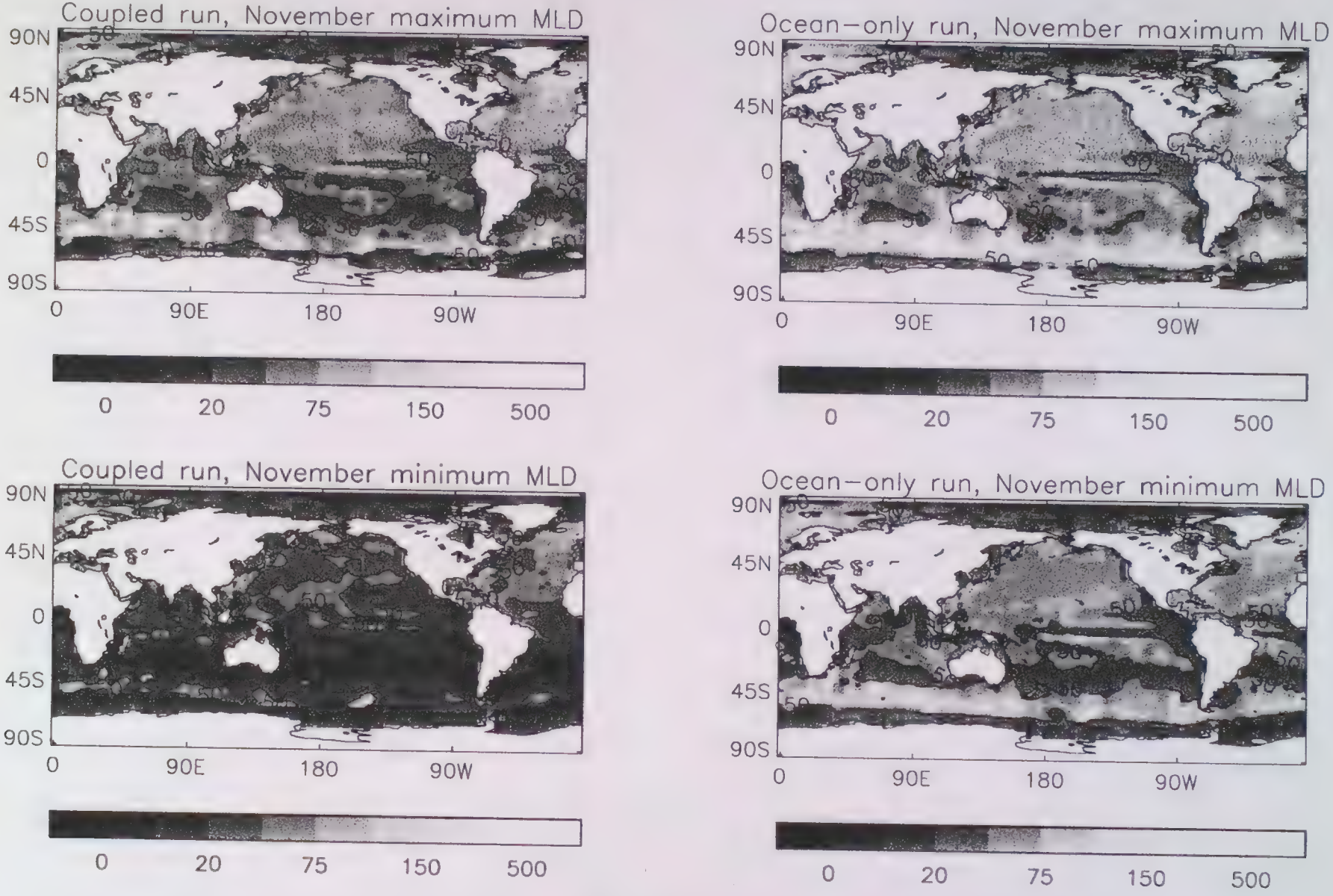

Figure 6. Maximum and minimum November mixed layer depths in metres. Maxima from the start of the second coupled phase (top left panel) and the end of the ocean only phase (top right panel). Minima from the start of the second coupled phase (bottom left panel) and the end of the ocean only phase (bottom right panel).

for quantities such as wind mixing energy which are physically positive. The best approach to take therefore with fluxes which must be positive would be to recalculate them with a scaling rather than a shift as shown in equation 2. But this carries the penalty that the range of values of f_{oc} is also scaled. Equation 1 is used in this study.

$$f_{oc} = f_{cp} * \frac{F_{decade}}{F_{year}} \tag{2}$$

The results show that the use of daily data greatly reduces the MLD discrepancy between the coupled and ocean only phases of the spinup. In particular the minimum depths during the daily ocean-only phase (not shown) resembled those from the coupled phase much more closely. Figure 7 shows annual mean values zonally averaged from the two coupled phases, and the two ocean only runs. It is clear that the daily forced ocean only run was much closer to the coupled runs than was the monthly forced run. This is particularly evident in the region 30°-60°S where the errors were greatest. Figure 8 shows how the global mean temperature changes with depth during the 2000 years of the ocean only stage. The large warming due to the deepened mixed layer in the monthly forced run (dashed line) can be seen at around levels 5-10. This warming is largely removed in the daily forced run (solid line). The two runs show very similar evolution of the deep ocean temperature. Further results from the spinup with the daily forced ocean only phase are described in detail in the next section.

The differences in diagnosed mixed layer depths discussed above are not necessarily indicative of large differences in the ocean state. Only small differences in the vertical profile of ocean temperature led to the noticeable differences observed in the MLDs. However, it is important to remove this source of drift from the recoupled phase. Also, for the purposes of the biology model, which makes use of diagnosed MLDs, it is important for them to be accurately modelled. Biological growth is sensitive to both the temperature and depth of the mixed layer, and hence the original (monthly forced) ocean only phase was not of acceptable quality for spinning up HadOCC. The daily forced run is of much better quality, justifying the expense of the extra processing required to force the ocean on a daily varying basis (this adds less than 10% to the computing time of the spinup).

6 Spinup results

The temperature drift observed on recoupling the monthly forced ocean-only phase is almost completely eliminated when daily forcing is used. Figure 9 shows this for the two recoupled phases. The mean drift after the daily forced rerun is of the order of 0.007°C/100 yrs, less than a quarter observed after the monthly forced ocean phase. This reduction in drift holds true for individual ocean basins except in the North Atlantic, where a slight positive drift is slightly increased. Figure 10 shows this for the South Pacific. SST errors relative to GISST, from a flux adjusted run beginning from the end of the second spinup are of the order of 0.17°C compared with 0.21°C in the first spinup experiment

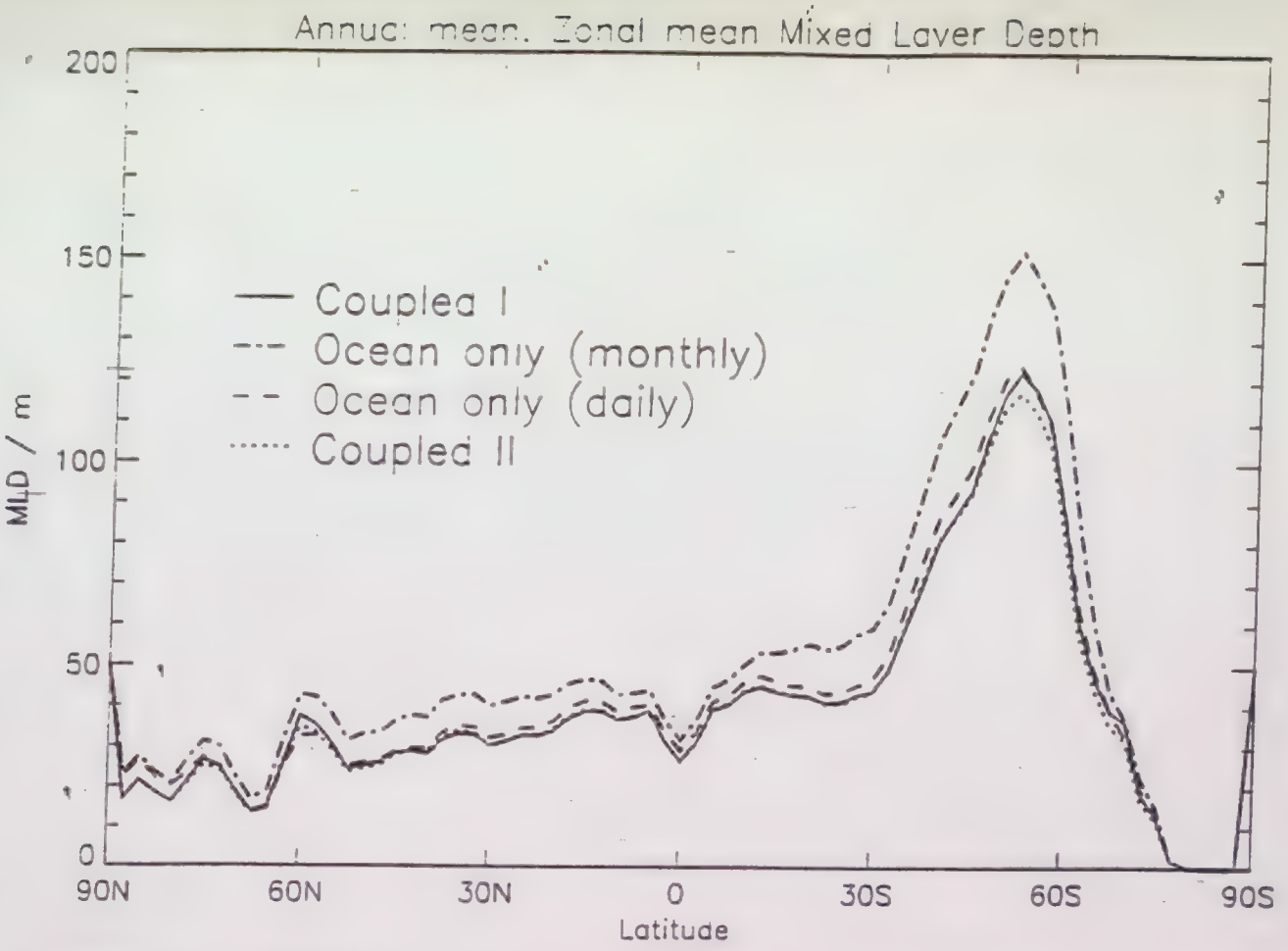

Figure 7. Annual mean zonal mean mixed layer depth for the two Haney forced coupled phases and the ocean only phase with monthly and daily forcing data.

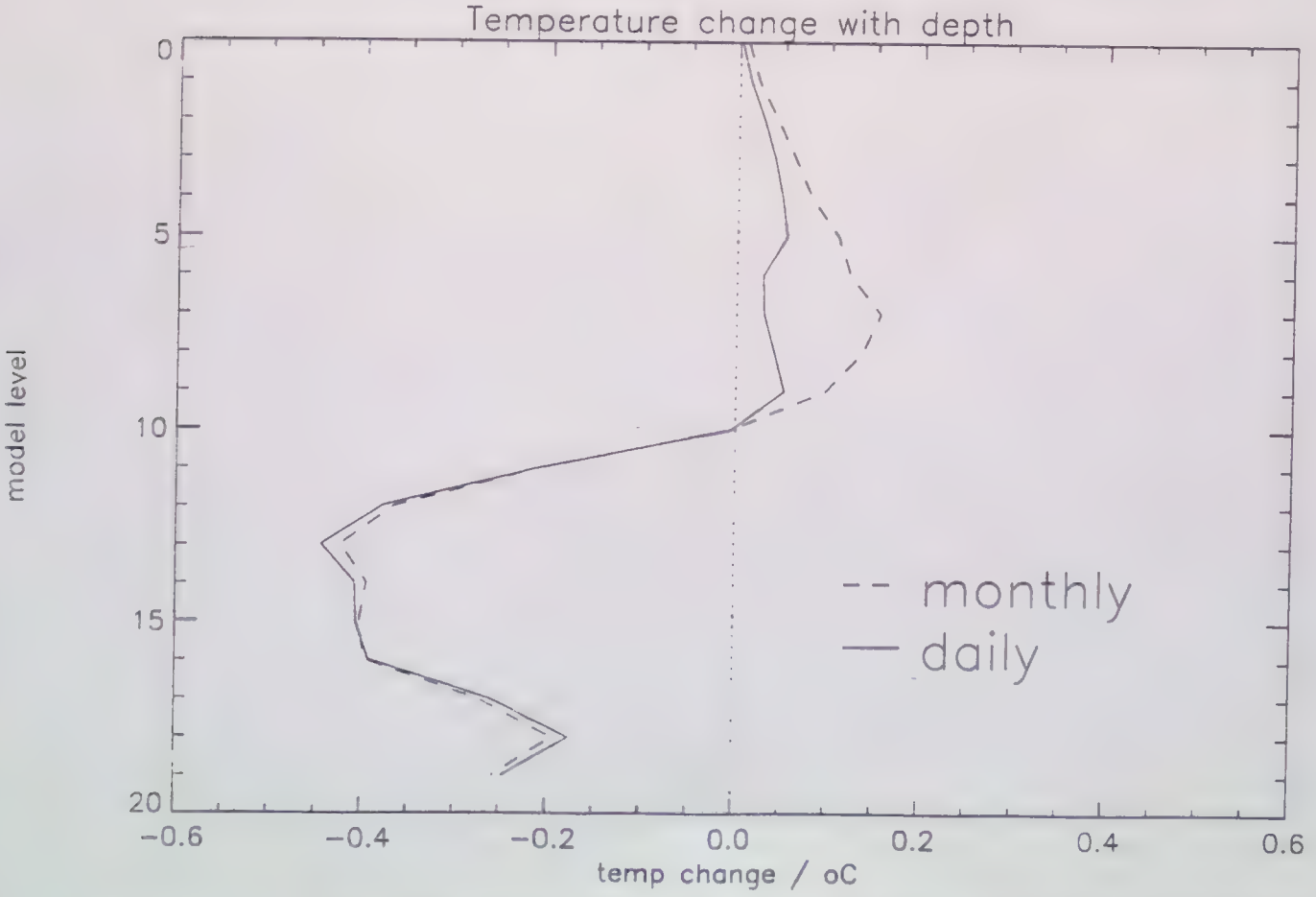

Figure 8. Global mean temperature change during the ocean only phases as a function of model level. Runs with monthly mean forcing (dashed line) and daily mean forcing (solid line).

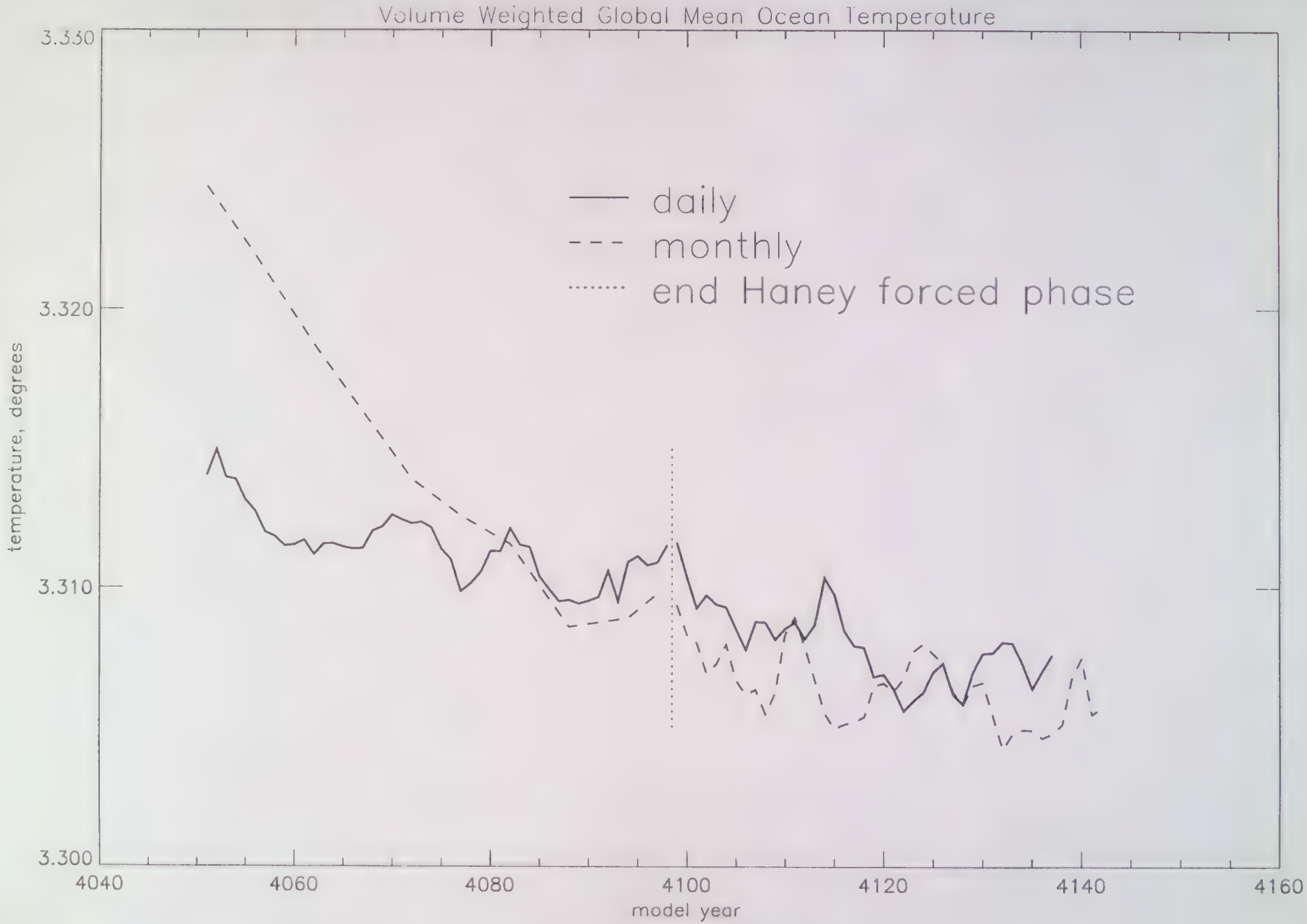

Figure 9. Global, volume weighted, mean temperature drift on recoupling after the ocean only phase with monthly mean forcing (dashed line) and daily mean forcing (solid line). Second Haney forced phase to the left of the dotted line, and the flux-adjusted phase to the right.

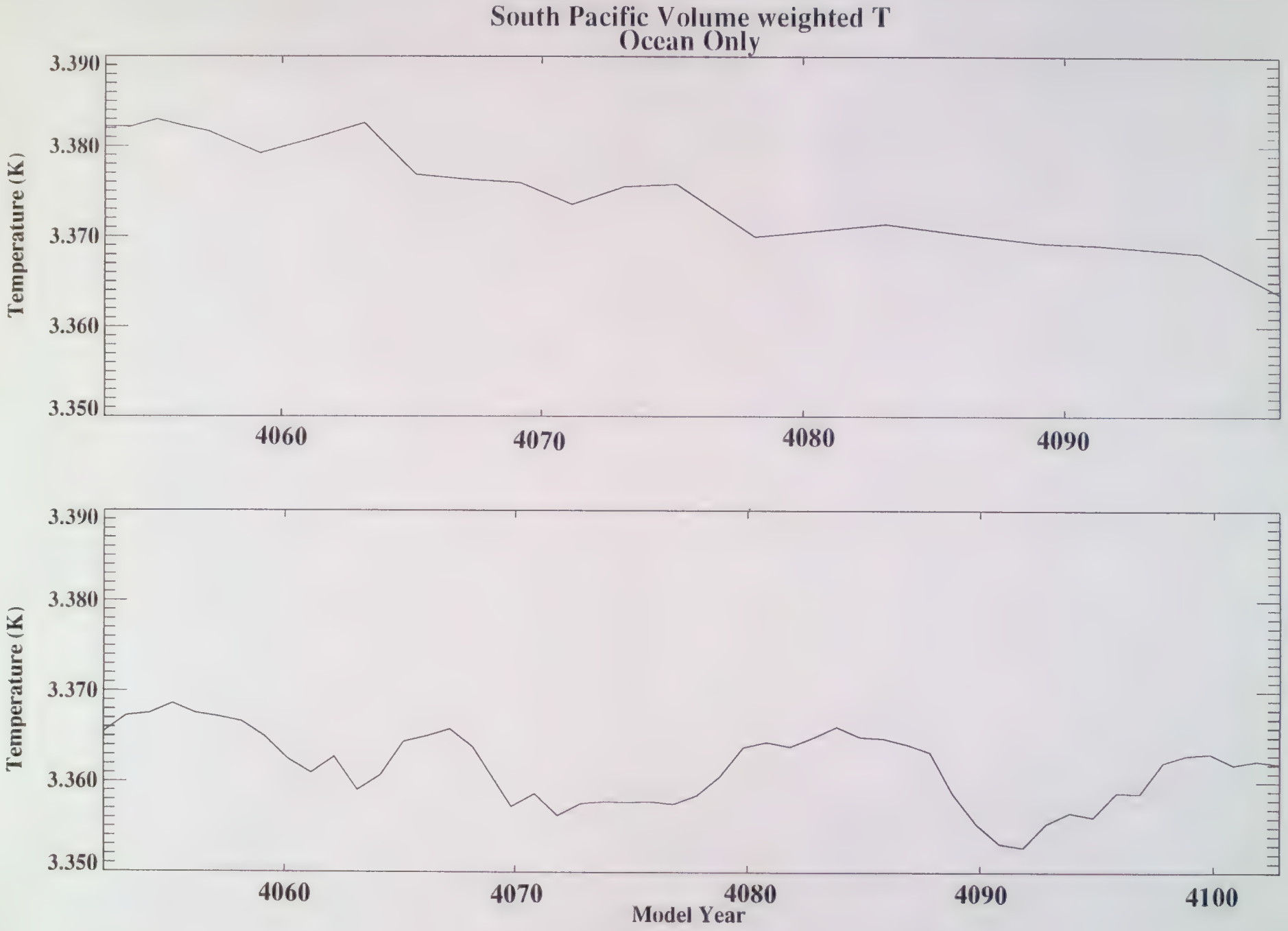

Figure 10. As figure 9, but for the South Pacific basin. Monthly mean forced run (top panel) and daily mean forced run (bottom panel).

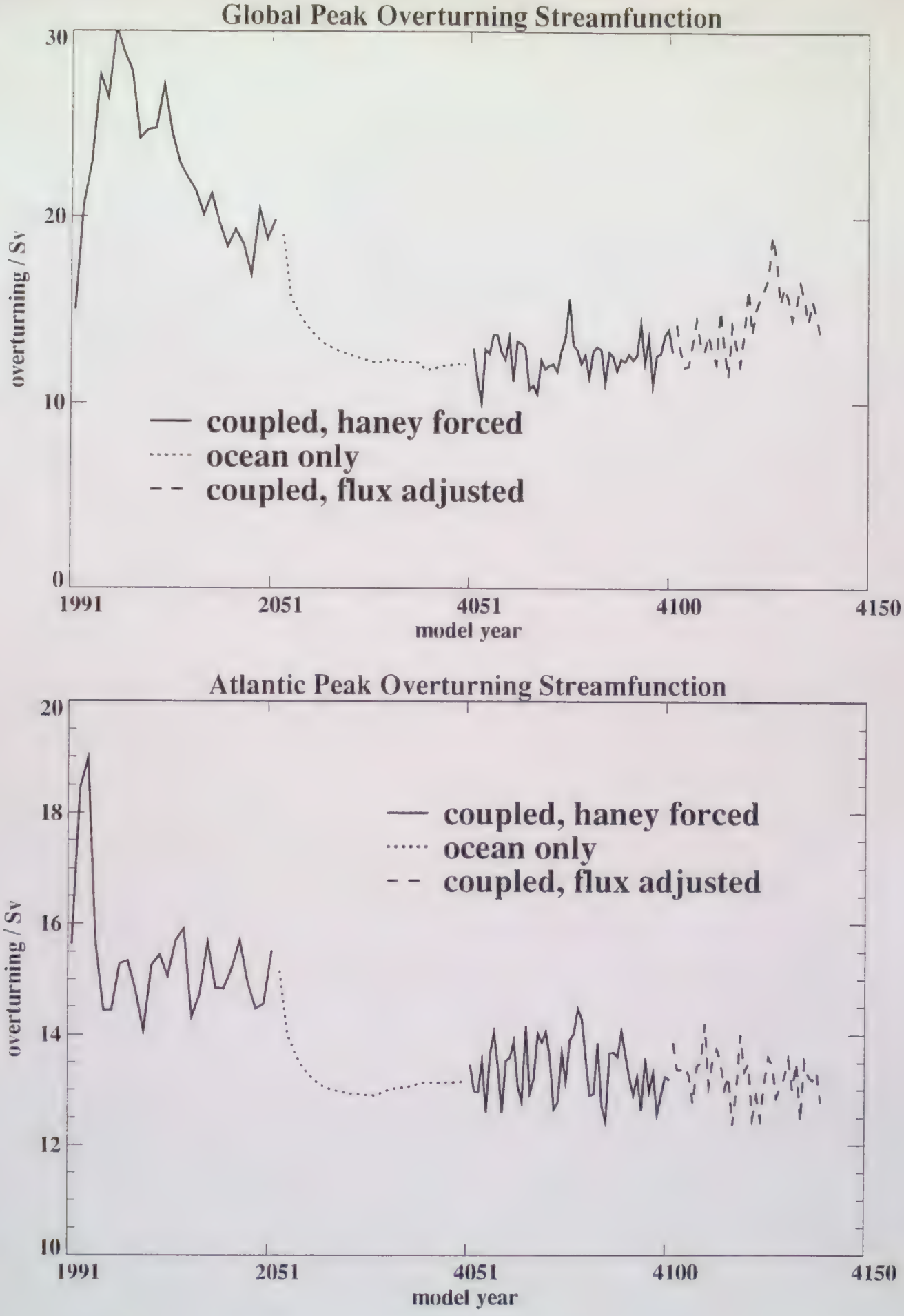

Figure 11. Timeseries of the magnitudes of annual mean meridional overturning. Global northward flow of bottom water (top panel) and Atlantic overturning associated with the formation of North Atlantic Deep Water (bottom panel).

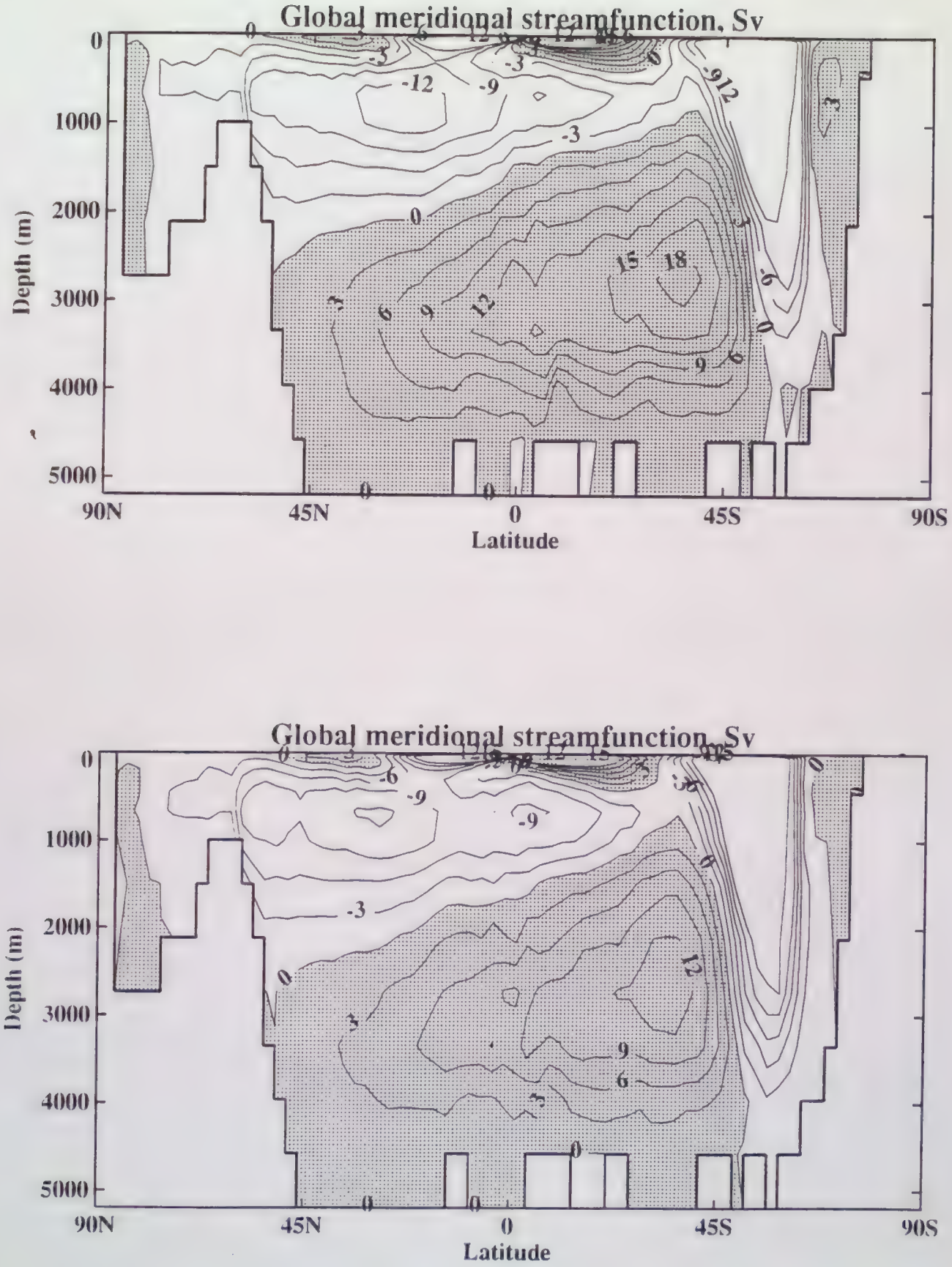

Figure 12. Decadal mean global meridional overturning streamfunction for the final decade of the first coupled phase (top panel) and the second coupled phase (bottom panel).

(with monthly forcing).

Figure 11 shows how peaks in the meridional overturning streamfunction evolve during the spinup. Values plotted are in Sverdrups and show the global northward flow of bottom water (top panel) and the Atlantic overturning associated with the formation of North Atlantic Deep Water (bottom panel). It can be seen that the magnitude of the circulation reaches a steady state in both cases. Figure 12 shows decadal mean global plots of the meridional streamfunction for the end of the first (top panel) and second (bottom panel) coupled phases. The final state is satisfactory given that observational data on actual values is sparse, although the Atlantic overturning may be a few Sverdrups higher (R. Wood, personal communication). Note that Gent-McWilliams eddy velocities are not included in these figures (if they were then peak values would be 1 to 2 Sv higher).

During the three phases of the spinup the ocean surface is Haney forced with a relaxation time of 2 weeks to keep it very close to climatological temperature and salinity values. However, during the flux adjusted run the ocean surface is free to drift. Figure 13 shows a timeseries of the global SST error relative to annual mean values from GISST. The error is small (about 0.2 °C) and shows little or no drift. Plots of individual years (not shown) show that the system has a (qualitatively) realistic inter-annual variability.

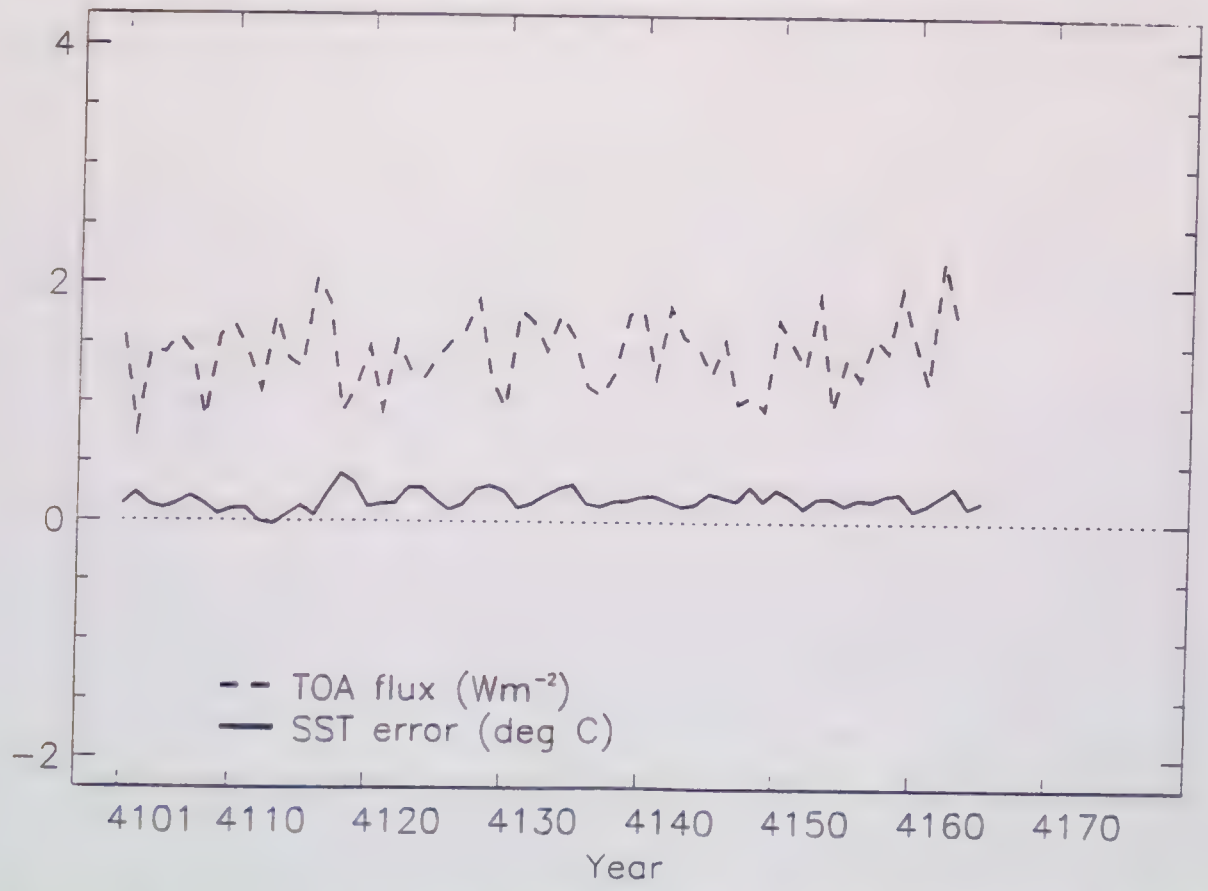

Figure 13. Timeseries of global annual mean sea surface temperature error relative to GISST climatology (solid line) and net radiation flux at the top of the atmosphere (dashed line) for the flux adjusted phase of the spinup.

Figure 13 also shows the net radiation flux at the top of the atmosphere (TOA). The global mean of 1.4 Wm^{-2} should ideally balance the global mean anomalous heat flux, which has a value of -2.1 Wm^{-2} in the final 10 years of the previous Haney forced coupled phase. The residual flux of -0.7 Wm^{-2} shows that the system is not in perfect equilibrium, but the error is small. The quality criteria specified in the PID for the TOA flux is +- 1.0 Wm^{-2}. There is no apparent drift in the TOA flux over the duration of the coupled phase.

7 Conclusions

The Coupled Climate-Carbon Cycle Project aims to include a fully interactive CO_2 component in climate change simulations, using models of the carbon cycle for both the ocean and terrestrial biosphere. A long spinup (of order 2000 years) is required due to the long equilibration timescales of the deep ocean and its carbon cycle fields. To perform this spinup using synchronous coupled running of the model would take several years real time based on current UKMO computer resources.

A periodically synchronous coupling technique has been developed which reduces the length of the spinup to about 2 months of real time. This comprises sixty years of coupled running, followed by 2000 years of ocean only running (with a 24 hour timestep rather than the normal 1 hour step), followed by a final 50 year re-coupled phase. The ocean only phase is forced by fluxes derived from the final decade of the first coupled phase, which are calculated so as to include the effects of sea ice because the ocean only phase is run with non-interactive sea ice.

Previous studies on this type of strategy have shown that such techniques can reproduce the results from synchronously coupled model runs, in much shorter computer time, and that the technique can be extended to work for GCMs as well as simple energy-balance "box" models. The new developments of our scheme are the accelerating of the ocean only phase by lengthening the timestep, and the performing of the ocean only phase in one long run rather than several short runs.

Due to non-linearities in the response of the model to forcing data, monthly mean fluxes did not produce the same mean response in the ocean phase as was seen in a subsequent coupled phase. Hence it was found to be necessary to force the ocean only phase with fields which included daily variability from the coupled run. The important point here is not just that the ocean state needs to be realistic during the ocean only phase (this could be achieved by retuning the ocean only model to respond to monthly mean forcing), but that there must be consistency between the coupled and ocean only phases. By ensuring that the ocean only phase sees forcing with the same frequency of variability as it would during the coupled phase, the two regimes are made consistent and the recoupling "shock" which manifests itself as a temperature drift, can be all but eliminated.

Results from the spinup have demonstrated that HadCM3L had reached close to an equilibrium state by the end of the second coupled phase. The temperature drift is small, and the magnitude of the overturning stream function is constant. This demonstrates that both the dynamics and thermodynamics of the ocean are stable.

8 Acknowledgements

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10 Appendix A.

HadCM2 (Hadley Centre Coupled Model, version 2) is described in depth in Johns (1996) and Johns et al. (1997). This appendix summarises the model set up and describes the important modifications to HadCM2 that were included in HadCM3 (and therefore HadCM3L).

10.1 HadCM2

The atmosphere and ocean components use a congruent rectangular grid (to facilitate coupling), with a resolution of 2.5° in latitude and 3.75° in longitude.

The atmosphere component equations are a quasi-hydrostatic version of the full primitive equations, with a representation of the Coriolis force. The prognostic variables are surface pressure, horizontal wind components, liquid water potential temperature, and total water mixing ratio. The model is formulated on an Arakawa "B" grid, with 19 vertical levels and a timestep of 30 minutes. Vertical velocity vanishes at the top and bottom coordinate surfaces. The integration scheme is split-explicit and 4th order accurate in space. The dynamical equations are formulated to conserve mass, energy, momentum, angular momentum and total water. Parameterisation processes include gravity wave drag, radiation, large scale cloud and precipitation processes, boundary layer physics and a single level soil hydrology scheme. Convection is calculated using a penetrative mass flux scheme with explicit downdraught representation.

The ocean component is based on the Bryan-Cox primitive equation model, with 20 vertical levels. Level thicknesses range from 10m near the surface in order to resolve the mixed layer, to 616m near the bottom of the ocean. The mixed layer uses an energy balance model, and negative surface buoyancy fluxes are convectively mixed down to a neutrally buoyant level. Wind stress contributes to turbulent kinetic energy which is assumed to decay exponentially with depth (with a decay length scale of 100m). Vertical diffusion is performed for both tracers and momentum, and a latitude-dependent horizontal diffusion is also applied. A mixing parameterisation is applied to simulate outflow from the Mediterranean Sea because there is no direct connection between the Mediterranean and the Atlantic in the model's topography.

The sea-ice component is a zero-layer thermodynamic model, on the ocean grid. Effects of brine rejection are included, and heat, water and momentum fluxes into the ocean are weighted by the leads fraction. The top level ocean

current is used to advect sea-ice.

The atmosphere and ocean components are run asynchronously for 1 day at a time. Mean fluxes of heat, momentum, water and wind-mixing energy are calculated from each atmosphere day and used to drive the subsequent ocean day. Conservation of fluxes is ensured by the congruence of the grids. It is possible to force the ocean surface in two other ways. Haney forcing can be applied, whereby the ocean surface temperature and salinity are relaxed to prescribed values. A typical relaxation coefficient is 163.76 $Wm^{-2}K^{-1}$ for temperature which corresponds to a relaxation time of about 2 weeks for a 50m mixed layer, and a coefficient for salinity which corresponds to the same relaxation time. Typically, the prescribed temperature and salinity values used are calculated every 5 days by linearly interpolating between monthly mean values of a climatological database. Secondly, flux adjustments can be applied. These are fixed heat and salinity fluxes derived from a Haney forced phase which are used to ensure that the mean model behaviour does not drift, whilst still allowing inter-annual variability. Typically flux adjustments are small in the global mean (a few Wm^{-2}) but can be large locally (several hundred Wm^{-2}).

10.2 HadCM3

HadCM3 uses a higher resolution ocean of 1.25 ° x 1.25 °. HadCM3L has the same ocean resolution as used for HadCM2.

The HadCM3 atmosphere component includes the Edwards-Slingo radiation scheme (Edwards and Slingo, 1996), the MOSES land surface scheme (Met Office Surface Exchange Scheme, Cox et al. 1998), improvements to gravity wave drag and the boundary layer schemes, the inclusion of convective momentum transport, the use of background aerosols in radiation, and a reduced value of the critical relative humidity (0.85 instead of 0.7) for cloud formation.

The HadCM3 ocean component includes a new vertical diffusion scheme due to Large (Large et al., 1994), an eddy parameterisation due to Gent and McWilliams (Gent and McWilliams, 1990) with a geographically varying coefficient (although a constant coefficient of 1000 cm^2s^{-1} was used in HadCM3L), improved Mediterranean outflow (where mixing is scaled to simulate a throughflow of 1 Sv), and a simplified sill overflow parameterisation for the Greenland-Scotland ridge.

11 Appendix B.

The forcing fluxes derived from the first coupled phase, which are then used to drive the ocean only phase are: penetrative solar heat flux, non-penetrative heat flux, wind mixing energy, wind stress (x and y components) and fresh water flux due to precipitation minus evaporation.

These are available from the UKMO Unified Model as the following diagnostics:
i) penetrative solar heat flux (SOL, stash code 161)
ii) non-penetrative heat flux (HTN)
iii) wind mixing energy (WME, stash code 152)
iv) wind stresses (Tau_X, Tau_Y, stash codes 150,151)
v) fresh water flux (PME)

HTN and PME are not the simple ocean diagnostics with stash codes 162 and 165 as might be expected. This is because the ocean only phase was run without any ice model, and so the effects of sea-ice had to be included in the atmospheric forcing fields that were applied. They were calculated in the following way:

$$HTN = HTN_{ocean} - O2I - CH_{30} + CH_0 - Q * S \tag{3}$$

where (stash codes in brackets):

HTN_{ocean} (30206) is HTN(162) weighted by the leads fraction to allow for some heat going into the ice and not the ocean.

O2I (143) is the ocean to ice heat flux.

CH_0 ($caryheat_0$, 142) and CH_{30} ($caryheat_{30}$, 30208) are heat fluxes associated with ice processes calculated before and after row calculations respectively.

S ($snowrate_{ocean}$, 30207) is snow rate where there is no ice (snow goes onto ice if there is any)

Q=333 520 JKg^{-1}, is the latent heat of fusion for snow melting in the sea.

$$PME = PME_{model} + R - CS * \frac{\Delta Z_1 . \rho_w}{sal_{ref}} \tag{4}$$

where:

PME_{model} (165) is precipitation minus evaporation

R (runoff, 166) is river runoff amount

CS (carysalt, 144) is a salinity flux associated with ice processes.

CS is scaled to transform it from a salinity flux to a fresh water flux using:

ΔZ_1, the depth of the top layer = 10m
ρ_w, the density of sea water = 1026 Kg m^{-3}
sal_{ref}, a reference salinity = 0.035

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